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Estimation of Net Physical Transport and Hydraulic Residence Times for a Coastal Plain Estuary Using Box Models

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ABSTRACT: A box model based on salinity distributions and freshwater inflow measurements was developed and used to estimate net non-tidal physical circulation and hydraulic residence times for Patuxent River estuary, Maryland, a tributary estuary of Chesapeake Bay. The box model relaxes the usual assumption that salinity is *at* steady-state, an important improvement over previous box model studies, yet it remains simple enough to have b a d appeal. Average 2-dimensional net non-tidal circulation and residence times for 1986–1995 are estimated and related *to* river flow and salt water inflow as estimated by the box model. *An* important result is that advective exchange *at the* estuary mouth was not correlated with Patuxent River flow, most likely due to effects of offshore salinity changes in Chesapeake Bay. The median residence time for freshwater entering at the head of the estuary was 68 d and decreased hyperbolically with increasing river flow to 30 d during high flow. Estimates of residence times for down-estuary points of origin showed that, from the head of the estuary to its mouth, control of flushing changed from primarily river flow to other factors regulating the intensity *of* gravitational circulation.

Introduction

The residence time of water is an important physical control on ecological processes in estuaries. For example, Nixon et al. (1996) showed that the fraction of nitrogen inputs subsequently exported from an estuary decreased as residence time increased. Other papers also cite ecological or geochemical effects of residence time in estuaries (e.g., Monbet 1992; Paucot and Wollast 1991; Muller et al. 1994), but the full potential *of* residence time as an explanatory variable in estuarine ecology has most likely not been realized because of the challenge of estimating it at the appropriate time and space scales. This paper presents a reasonably simple model that has been used to estimate physical transport and spatially resolved residence times for Patuxent River estuary, Maryland, at a monthly time interval.

Because residence time has been defined in a variety of ways (e.g., Miller and McPherson 1991), one definition will be adopted here and will apply to all subsequent references to residence time. Hydraulic residence time is defined as the mean amount *of* time a parcel of water remains in the estuary once it enters. Equivalently, residence time is the time required to reduce the total mass in the estuary of an introduced pulse of a conservative material by e^{-1}, or 36.8% of the *original* total mass. Hydraulic residence time refers to the residence time of water, as opposed to particles or dissolved or suspended materials in the water.

A variety of methods have been developed to estimate residence times for estuaries. Hydrodynamic simulation methods are well developed and provide the most information (e.g., residence times, water parcel age, transit times, transport processes). Vallino and Hopkinson (1998), using several models (including a salt balance model) and field studies, provide a recent example of what is possible, albeit in a I-dimensional circulation. Unfortunately, these complex models are often beyond the capability of many ecologists and other environmental scientists and managers. This is espe-

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cially true where 2-dimensional or 3-dimensional representations of the circulation are required. Salt-balance approaches are relatively simple tools for ecologists that can yield useful ecological insights (Hagy 1996).

One advantage of salt-balance models is that they have been applied to estuaries for some time and are reasonably well known. Pritchard (1960) estimated residence time for Chincoteague Bay, Maryland based on both intertidal volume and salt balance, providing estimates of flushing time and predicting the effect of closing an ocean inlet on mean salinity. The popular but crude fraction-of-freshwater method, reviewed by Dyer (1973), also provides a simple method of estimating residence times for a whole estuary. Two major limitations of these methods are that spatially resolved residence times are not possible and that salinity must be assumed constant This makes these metho most suitable to estimating long-term average residence times. Pilson (1985) used a variation of this method to estimate residence times for Narragansett Bay under a variety of flow conditions, statistically relating residence times to river flow rate. Recognizing and estimating the dependence of residence time on river flow was an important step forward, but the poor compliance with the steady-state assumption for salinity may have introduced error. Asselin and Spaulding (1993) validated salt-balance based estimates with tracer release experiments to estimate residence times at different river flow levels. Such validation studies, also used by Vallino and Hopkinson (1998), are useful in combination with modeling, but aren't always practical due to cost or logistical considerations. In addition, dye studies cannot be used for retrospective analysis. Miller and McPherson (1991) made an important advance in application of salt-balance methods to estimate residence time. They used a single layer case of the box model approach of Officer (1980) to estimate tidal dispersion in a shallow estuary, and then estimated residence times by simulation. This approach provided the advantage of spatially resolved estimates of residence times, including the potential to estimate residence times defined in a variety of ways. Their assumption that tidal dispersion at any point in the estuary was independent of river flow allowed them to predict salinity distributions using the model, providing a convenient means of validation, even for a retrospective study. Unfortunately, this approach cannot be applied to a two-layer estuary where gravitational circulation is not likely a constant Miller and McPherson (1991) addressed the problem of satisfying the steady-state assumption by estimating steady-state salinity profiles.

We estimate residence times for Patuxent River estuary as a function of river flow. We use a box model and dynamic simulation approach similar to Miller and McPherson (1991), but with several important elaborations. The box model utilizes a mixed one-layer, two-layer box model (Pritchard 1969; Officer 1980), as is appropriate for a partially stratified estuary such as the Patuxent This permits estimation of residence times as a function of both flow and point of origin (sensu Miller and McPherson 1991). Our model also accounts for both seasonal changes in salinity and multiple freshwater sources, advancements in the application of box models that have not been used elsewhere.

Study Site

The Patuxent River estuary, Maryland, is an ideal site for this study because of the availability of a long-term and spatially resolved record of salinity collected by the Chesapeake Bay Water Quality Monitoring Program (Fig. 1; Environmental Protection Agency Chesapeake Bay Program Office). The estuary is approximately 65 km in length, has a mean-low-water estuarine volume of 577×10^6 m^3 and a surface area of 126×10^6 m^2. Over the most seaward 45 km, the estuary averages 2.2 km in width and 6.0 m in depth (Cronin and Pritchard 1975). The tide has a mean range of 0.4 m near Solomons, Maryland, and increases landward to near 0.8 m (Boicourt and Sanford 1988). The water column is vertically mixed in the upper estuary and seasonally stratified in the lower estuary. The area of the drainage basin above the fall line at Bowie, Maryland, is 901 km^2, accounting for 39% of the total watershed area (Environmental Protection Agency 1998). Fall line freshwater discharge averaged 9.6 m^3 s^{-1} during 1986–1995 (United States Geological Survey 1985–1996).

Methods

The Box Model

This study used a modification of the box model approach of Officer (1980) to estimate advective and non-advective ((diffusion and dispersion) transport. This method uses observed distributions of salt, as a conservative tracer, known freshwater inputs, and estuarine geometry to estimate estuarine exchange coefficients, which can then be used in other calculations. Two important assumptions of Officer (1980) were relaxed, specifically, the assumptions that salinity is at steady-state and that all freshwater enters at the head of the estuary. To accommodate these changes, the time rate of change of salinity for each month was estimated from the time series of salinity data. Freshwater inputs to each segment of the model were estimated.

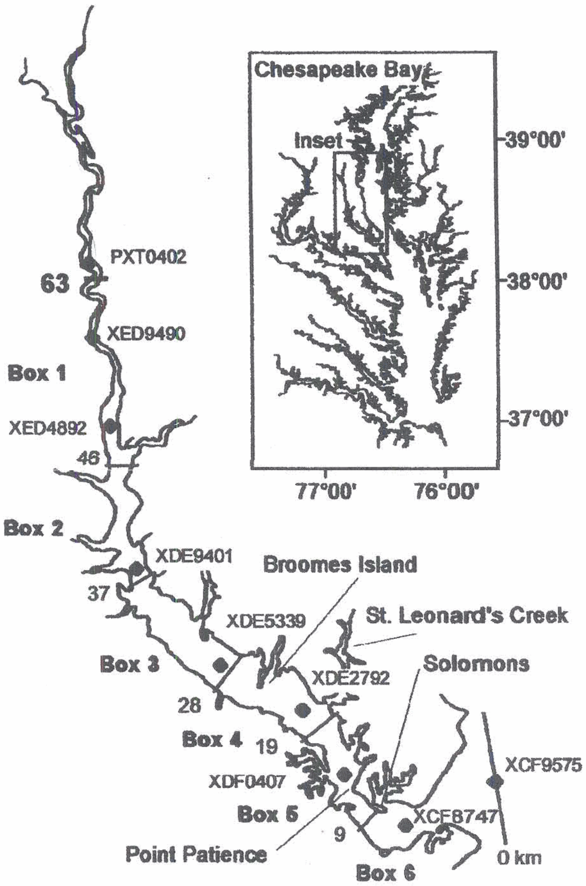

Fig. 1. Map of Patuxent River, Maryland showing the location within the Chesapeake Bay system, the locations of box boundaries and the locations of Chesapeake Bay Water Quality Monitoring Program stations. Numbers next to box boundaries indicate channel distance from the estuary mouth in kilometers.

The details of these calculations are provided below under appropriate headings.

We can use box models to estimate advective and non-advective exchanges between a volume and the surrounding environment (Fig. 2, upper panel) by solving a system of linear equations describing the salt and water balance. For the generic volume in Fig. 2, the equation for salt balance is

$$V\frac{ds_{in}}{dt} = Q_{in}s_{out} - Q_{out}s_{in} + E(s_{out} - s_{in}) \quad (1)$$

where the terms are defined as follows: Q_{in} = new water advection into the control volume; Q_{out} = advective transport out of the control volume; E = non-advective exchange between the control volume and the surrounding environment; s_{in} = salinity inside the control volume; s_{out} = salinity outside the control volume; and V = volume of the control

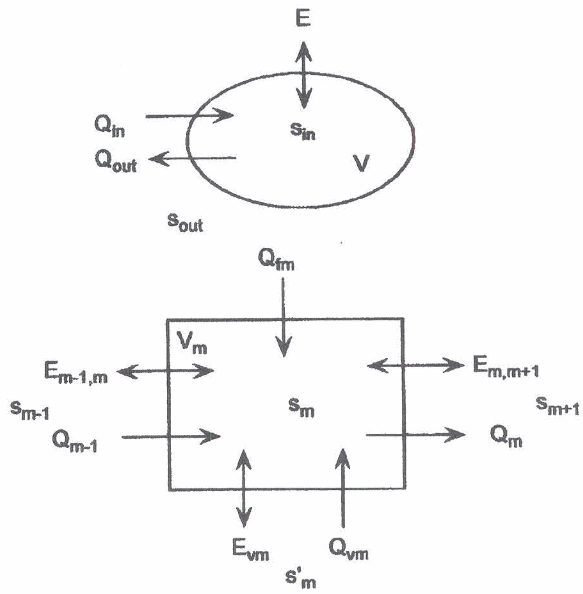

Fig. 2. A diagram depicting the salt and water exchanges for a generic control volume (upper panel) and for any box in a mixed one-layer and two-layer estuarine box model (lower panel). For any particular box in the model, some of the exchanges may have zero or negative values.

volume. The volume is assumed to remain constant at the time scales of interest (i.e., sub-tidal or >1 wk). Because the water balance is $Q_{in} = Q_{out}$, Eq. 1 reduces to:

$$V\frac{ds_{in}}{dt} = (Q_{in} + E)(s_{out} - s_{in}) \quad (2)$$

This general case can be applied to each box in a multi-box model for an estuary. The possible exchanges include horizontal advective and non-advective exchanges in two directions, vertical advective and non-advective exchanges, and freshwater input (Fig. 2, lower panel). The salt balance is

$$V_m\frac{ds_m}{dt} = Q_{m-1}s_{m-1} + Q_{vm}s'_m - Q_m s_m + E_{vm}(s'_m - s_m) + [E_{m-1,m}(s_{m-1} - s_m) + E_{m,m+1}(s_{m+1} - s_m)] \quad (3)$$

and the water balance is

$$Q_m = Q_{m-1} + Q_{vm} + Q_{fm} \quad (4)$$

where the terms are defined as follows: Q_m = advective transport to the down-estuary box; Q_{m-1} = advective transport from the up-estuary box; Q_{vm} = vertical advective transport into the box; Q_{fm} = freshwater input into the box; $E_{m-1,m}$ = non-advective exchange with the up-estuary box; $E_{m,m+1}$ = non-advective exchange with the down-estuary

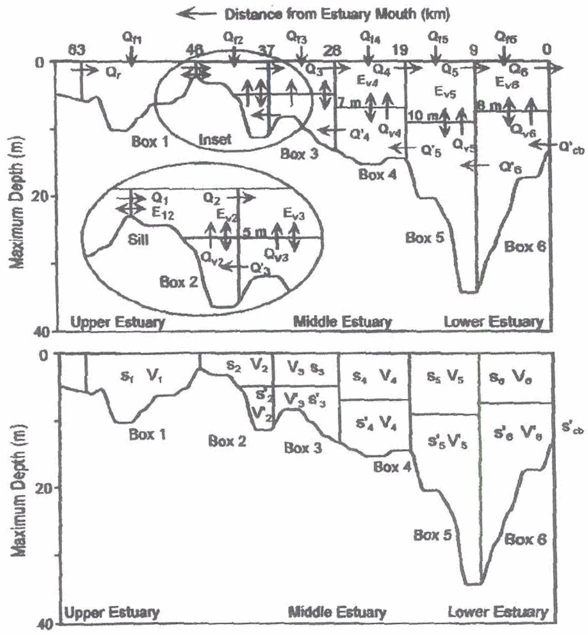

Fig. 3. Schematic diagrams of the box model structure showing the box boundaries, the exchange coefficients that were estimated, and the model inputs. The estimated exchanges include seaward advection (Q_m), landward advection (Q'_m), vertical advection (Q_{vm}), vertical diffusive exchange (E_{vm}), and horizontal dispersion (Q„ ,). Inputs included the volume in each box (V_m and V'_m), concentration of salt for each box (s_m or s'_m), river flow (Q_r), the input of freshwater to each box (Q_{fm}), and the salinity at the seaward boundary (s'_{cb}). The sill noted at the boundary of box 1 and box 2 terminates the landward flow in the bottom layer, forcing the water into the surface layer.

box; E_{vm} = vertical non-advective exchange; s_m = salinity in the box; s_{m-1} = salinity in the up-estuary box; s_{m+1} = salinity in the down-estuary box; s'_m = salinity in the vertically adjacent box; and V_m = volume of the box

If all horizontal and vertical advective and non-advective terms in Eq. 3 were included, there would be 32 exchange coefficients to determine from these 22 equations and the system would be under-determined and unsolvable. In order to make the system solvable, we have assumed that horizontal non-advective exchange (longitudinal dispersion) is negligible compared to horizontal advective exchange in the region of the estuary where there is a well-developed two-layer gravitational circulation (boxes 2–6; Fig. 3). This assumption is discussed later, where it is shown to be quite reasonable for the Patuxent River. It eliminates 10 exchange coefficients and allows ready solution of the system of equations. Simplifying the model, the salt balance equation (Eq. 3) for surface layer box 4 reduces to

$$V\frac{ds_4}{dt} = Q_3 s_3 + Q_{v4}s'_4 - Q_4 s_4 + E_{v4}(s'_4 - s_4) \quad (5)$$

Box 1 represents a one-layer transition area between the river and the estuary. Advective and non-advective exchanges between the bottom layer of box 2 and box 1 were assumed to be zero due to the presence of a sill (Fig. 8). In the absence of this sill, it would be necessary to assume a relationship between the surface and bottom layer exchanges at the transition from one-layer to two-layer regions of the model as described by Officer (1980). Box 1 has non-zero salinity due to dispersion but not through two-layer circulation. Equation 3 for box I, reduces to

$$V\frac{ds_1}{dt} = E_{12}(s_2 - s_1) - Q_1 s_1 \quad (6)$$

Our box model structure is extended beyond Officer's (1980) equations to permit time-variable salinity and inputs of freshwater into each box. For example, Officer's Eq. (49) for seaward advection becomes

$$Q_m = \frac{\left(s'_{m+1}\left[\sum_{j=1}^{m} Q_{fj} + Q_r\right] + \left[\sum_{j=1}^{m} V_j\frac{ds_j}{dt} + \sum_{j=2}^{m} V'_j\frac{ds'_j}{dt}\right]\right)}{s'_{m+1} - s_m} \quad (7)$$

where Q_r is the river discharge at the head of the estuary and Q_{fj} is the indexed value of Q_{fm}, defined as above. V_j is the volume of bottom layer box j and the other terms are as defined above. Officer's Eq. (44), which yields landward advection, becomes

$$Q'_{m+1} = Q_m - \sum_{j=1}^{m} Q_{fj} + Q_r \quad (8)$$

Equation 7 contains additional terms on the right-hand-side in the numerator describing the change in salinity through time. Additional summation terms appear in both Eqs. 7 and 8 to account for freshwater inputs to each segment of the estuary.

The box model equations can be solved using only two equations at a time. This allows closed expressions for the model solution to be derived, averting the need for a matrix approach. The model can be computed using a spreadsheet or simple computer program (we used the latter). For example, vertical advective exchange can be calculated by

$$Q_{vm} = Q'_{m+1} - Q'_m \quad (9)$$

an expression that arises directly from the water balance equation for any bottom layer box Vertical non-advective exchange (E_{vm}) is obtained by solving Eq. 5 and takes the form

$$E_{vm} = \frac{\left(V_m \frac{ds_m}{dt} + Q_m s_m - Q_{m-1} s_{m-1} - Q_{vm} s'_m\right)}{s'_m - s_m} \qquad (10)$$

Similar expressions may be derived for the remaining exchange coefficients, but they are not presented here for brevity.

Box Model Data Requirements

The above equations require four types of inputs. These are box volumes (V_m and V'_m), freshwater input rates (Q, and %), salinity distributions (s_m and s'_m), and rates of salinity change (ds_m/dt and ds'_m/dt). Box volumes were obtained from Cronin and Pritchard (1975). Freshwater inputs (Q_{fm}), salinity distributions, and rates of salinity change were estimated as described below.

Water Budget

Estimates of freshwater inputs were made primarily on the basis of daily Patuxent River discharge measurements at the United States Geological Survey gauging station at Bowie, Maryland. This provided a good estimate of freshwater inputs from 39% of the watershed.

Water inputs from ungauged areas were estimated by calculating the water yield per area of the gauged watershed, then multiplying a fraction of that by the area of ungauged watershed. The fraction, L, in Eq. 11 reflects the fact that the water yield per area of ungauged watershed may not be exactly equal to the water yield of the gauged watershed due to hydrological differences caused by topography, geology, and land-use (Environmental Protection Agency 1998). Water yield for the ungauged watershed was calculated for each month using

$$Y = L\frac{k(Q_G + Q_R + Q_{MW})}{A} \qquad (11)$$

where Y is the water yield per area of watershed, Q_G is the measured discharge rate at the Bowie, Maryland gauge, Q_R is the net storage rate of water in reservoirs, Q_{MW} is the rate of water withdrawal for municipal use, A is the watershed area above the gauge, and k scales to appropriate units. Monthly average values for Q_R were obtained from a record of month-end water levels in the Triadelphia and T. Howard Duckett reservoirs (United States Geological Survey 1985–1996). Values for Q_{MW} were obtained from monthly municipal water withdrawal records (United States Geological Survey 1985–1996).

An estimate of the lower watershed yield relative to that of the upper watershed (L in Eq. 11) was obtained by comparing monthly averaged water yields for 1989, 1992, and 1993 for the gauged portion of the Patuxent watershed with concurrent yields from the Killpeck Creek and Hunting Creek watersheds. These small gauged watersheds are in the lower Patuxent River watershed.

Estimates of direct water inputs to the water surface were made using precipitation data for the Lower Southern District and pan evaporation measurements for Upper Marlboro, Maryland (National Oceanic and Atmospheric Administration 1986–1996). Evaporation rates were averaged by month across years because the data were relatively sparse and interannual differences were small. Water surface areas were obtained from Cronin and Pritchard (1975).

Freshwater input to each segment of the estuary ($F = R + P - E$) was calculated as the sum of runoff from the watershed (R) and direct precipitation to the water surface (P) minus evaporation from the water surface (E). Since the gauged flow includes groundwater inputs above the fall line, these inputs were implicitly included in the input estimates for ungauged areas, however, this is an area of some uncertainty. Each of these rates were multiplied by either watershed area in the case of R, or water surface area in the case of P and E to obtain the monthly freshwater input rate for each box.

Salinity Distributions and Changes in Salinity

Salinity data were obtained at 3 m depth intervals at 9-stations located along the main channel of the estuary using a variety of multi-probe (i.e., CTD) instruments (Environmental Protection Agency 1992; Fig. 1). Sampling occurred monthly during December through February and biweekly otherwise. To calculate the volume mean salinity (sensu Pritchard 1960) for each box, the data were interpolated using a quadrant-search linear interpolation algorithm adapted from Bahner et al. (1991) and Reynolds and Bahner (1989). The interpolated grid has 477 cells, each 1.85 km (1 nautical mile) in length, 1 m in vertical thickness, and extending the width of the estuary. The volume of each grid cell at mean low tide was obtained from Cronin and Pritchard (1975). Contour plots of the interpolated data were used to scan for poor interpolation or data problems. Mean salinity, weighted by cell volume, for each of the model boxes was calculated from the gridded data for each cruise in the time series. The time rate of change of salinity (ds_m/dt and ds'_m/dt) for each month calculated by difference from the time series of mean salinity.

Residence Time Calculations

Residence time calculations were made using the box model-derived estimates of the net (nontidal) circulation in a numerical simulation exactly in the manner of Miller and McPherson (1991). This simulation follows movement of a conservative, dissolved tracer material introduced in a single pulse. During the simulation, all water exchanged at the mouth of the estuary and all river inputs had no tracer. The residence time was calculated as the time required to reduce the mass of tracer in the entire estuary to e^{-1} times the initial mass. Exchange coefficients were held constant through the simulation so that estimated residence times could be unambiguously related to initial conditions at the time of pulse introduction. Thus, a residence time of 70 d might be computed even though the circulation regime never persists unchanged for 70 d. At each time step in the simulation, the change in amount of tracer present in each surface layer box was calculated as

$$V_m \frac{dc_m}{dt} = Q_{m-1} c_{m-1} + Q_{vm} c'_m + E_{vm}(c'_m - c_m) + E_{m-1,m}(c_{m-1} - c_m) - E_{m,m+1}(c_m - c_{m+1}) - Q_m c_m \tag{12}$$

with appropriately defined exchange coefficients estimated from measured salinity distributions as described above. The same expression for any bottom layer box is

$$V'_m \frac{dc'_m}{dt} = Q'_{m+1} c'_{m+1} - Q_{vm} c'_m - Q'_m c'_m - E_{vm}(c'_m - c_m) \tag{13}$$

In Eqs. 12 and 13, c_m, c_{m-1}, and c_{m+1} are the tracer concentrations in surface layer box m, $m - 1$, and $m + 1$, respectively. Following convention, c'_m and c'_{m-1} are identically defined, but for bottom layer boxes. Residence time calculations utilized a time step of 1 h and Euler integration, which was sufficient to reduce error to negligible levels according to test simulations run using considerably shorter time steps.

Simulations were initiated with a unit concentration of tracer in one or more boxes and zero concentration in all other boxes. It was assumed that once the tracer left the Patuxent River estuary, none of it returned (i.e., new water has concentration = 0). The relatively large volume and strong circulation of Chesapeake Bay, the body into which the Patuxent River estuary discharges, justifies this assumption. Sanford et al. (1992) describe a method for relaxing this assumption. Simulations were terminated when the total mass of tracer material remaining in the entire estuary was reduced by e^{-1} Residence time, T, is the time t where

$$\left(\sum_{m=1}^{6} V_m c_m + \sum_{m=2}^{6} V'_m c'_m\right)_{t=T} = e^{-1}\left(\sum_{m=1}^{6} V_m c_m + \sum_{m=2}^{6} V'_m c'_m\right)_{t=0} \tag{14}$$

Since the tracer was added in a pulse only at the beginning of the simulation (i.e., at $t = 0$), the residence times are called pulse residence times. Several types of residence times were defined according to the box or h e s into which the initial pulse was introduced.

If the simulated tracer was introduced only at the head of the estuary (e.g., box 1), the estimated residence time was defined as the pulse residence time for freshwater, or PRT_F. This calculation is analogous to the freshwater replacement time calculated using the fraction-of-freshwater method (Dyer 1973), but our calculation does not assume steady-state salinity. If the substance was introduced uniformly throughout the estuary, the estuarine residence time, or ERT, was obtained. Other pulse residence times, PRT_m, were calculated by introducing the simulated conservative substance into each surface layer box m of the model. This simulation method and the definitions of the residence time terminology are adapted from Miller and McPherson (1991). Residence times were calculated for each month of the 1986–1995 average year and for each individual month through the same period.

For comparison, the relatively well known fraction of freshwater method (Dyer 197.3) was used to calculated freshwater replacement time (*FRT*), a steady-state estimator of residence time. *FRT* was calculated as

$$FRT = \left(\frac{S_s - S_e}{S_s}\right)\frac{V}{Q_f} \tag{15}$$

where S_s is the salinity at the seaward margin of the estuary, S_e is the average salinity within the estuary, V is the total volume of the estuary, and Q_f is the total freshwater input

Hyperbolic functions were used to predict residence times as a function of both river flow and seawater inflow. This multiple regression model has the form

$$T = a + \frac{5.77(10^8)}{86400(bQ_r + cQ'_{cb})} \tag{16}$$

The constant in the numerator is the volume of Patuxent estuary (m^3), while 86,400 is the number

TABLE 1. Inputs to the 1986–1995 mean water budget for Patuxent River. Water yield for the upper watershed (A) was calculated according to Eq. 11. Water yield for the lower watershed was assumed to be 70% of A. Precipitation (B) is from National Oceanic and Atmospheric Administration (1987–1996). Evaporation (C) is average evaporation by month from National Oceanic and Atmospheric Administration (1987–1996) with December through March values inferred from the April through November observations. Ungauged freshwater inputs were calculated as A(0.70)E + F(B − C). The freshwater input to box 1 also includes the gauged flow (D) at Bowie, Maryland.

Month	A. Upper Runoff (mm d^{-1})	B. Precipitation (mm d^{-1})	C. Evaporation (mm d^{-1})	D. Flow-Gauged (m^3 s^{-1})	Total Freshwater Inputs (m^3 s^{-1})					
					Box 1	Box 2	Box 3	Box 4	Box 5	Box 6
Jan	1.40	2.87	0.40	12.2	22.2	1.7	1.9	1.9	1.6	1.2
Feb	1.34	2.48	0.40	11.3	20.9	1.6	1.7	1.7	1.4	1.1
Mar	1.91	3.74	1.64	17.1	30.7	2.0	2.2	2.1	1.8	1.4
Apr	1.44	2.56	3.66	13.3	23.3	1.0	0.9	0.7	0.6	0.3
May	1.44	3.10	4.46	12.9	22.9	0.9	0.8	0.7	0.5	0.3
Jun	0.87	2.95	5.36	7.5	13.4	0.2	0.0	−0.1	−0.1	−0.2
Jul	0.69	3.6	5.32	5.8	10.5	0.2	0.1	0.0	0.0	−0.1
Aug	0.57	3.08	4.36	5.2	9.1	0.4	0.1	0.0	0.0	−0.1
Sep	0.57	3.49	3.39	4.9	8.9	0.5	0.5	0.5	0.4	0.3
Oct	0.64	2.64	2.62	5.9	10.4	0.5	0.5	0.5	0.4	0.3
Nov	1.08	2.80	1.73	8.7	16.4	1.1	1.2	1.2	1.0	0.7
Dec	1.26	2.80	0.82	10.3	19.3	1.5	1.6	1.6	1.3	1.1
E. Watershed Area (km^2)—ungauged					867	104	102	94	78	54
F. Water Surface Area (km^2)					7	18	26	28	24	22

of seconds m 1 d. These values make b and c unitless quantities. Univariate models involving only Q_r or Q_{cb} were also fitted to the PRT_F and ERT estimates to simplify graphical presentation and to illustrate goodness-of-fit, since the meaning of the coefficient of determination (r^2) is uncertain for a non-linear model (Kvalseth 1985). The model relating PRT_r to Q_r was fitted using iteratively reweighted least squares (IRLS) regression (SAS Institute, Inc. 1990), a regression technique that is resistant to outliers.

Results

Water Budget

For the 1986–1995 period, water was retained in reservoirs from November through May at a mean rate of 0.67 m^3 s^{-1} and released from June through October at a mean rate of 0.93 m^3 s^{-1}. Gauged flows at Bowie, Maryland, were reduced by an average diversion of 1.89 m^3 s^{-1} of water to municipal water supplies at Laurel, Maryland. Correcting for these effects according to Eq. 11, and making calculations as described above, freshwater inputs from the upper watershed to box 1 (Fig. 3) were estimated (Table 1). Freshwater inputs for the 1986–1995 average year ranged kom a high of 47.6 m^3 s^{-1} in March to a low of 11.4 m^3 s^{-1} in August.

The comparison of water yield from the Killpeck Creek and Hunting Creek watersheds with the upper Patuxent watershed showed that the smaller watersheds delivered 27% less yield than the upper Patuxent watershed. These small gauged watersheds are the only direct measurements of water yield from the lower Patuxent watershed, which has lower topographic relief, more forest cover, and less urban areas than the upper watershed. To evaluate the above figure as an estimate of the water yield for the lower watershed as compared to the upper (L in Eq. 11), we compared the long-term mean water yield of the lower estuary estimated by the Chesapeake Bay Program Watershed Model to our calculations for the gauged upper watershed (Environmental Protection Agency 1998; Linker et al. 1999). This comparison found that the average water yield for the lower watershed was 69% of that for the upper watershed. Thus, we assumed that the water yield of the lower watershed was 70% of that for the upper watershed for the purpose of water budget calculations.

Freshwater inputs to the head of the estuary (box 1; Fig. 1) dominated the water budget throughout the year (Table 1). During winter and spring, this input was 20–30 m^3 s^{-1} and contributed about 75% of the total inputs. Through summer, it dropped to 10–15 m^3 s^{-1} but contributed up to 100% of the total freshwater input During June and July, evaporation from the broad lower estuary exceeded direct precipitation plus diffuse runoff into the lower estuary (Table 1).

Total freshwater input during the 1986–1995 average year ranged from 9.3 m^3 s^{-1} in August to 40.2 m^3 s^{-1} in March (Table 1). For individual months during 1986–1996, the range in freshwater input was 0.6 m^3 s^{-1} in August 1987 to 88.7 m^3 s^{-1} during March 1994 floods.

Long-term Mean Seasonal Circulation Patterns

Estimates of the advective and non-advective exchange coefficients were made for each month of

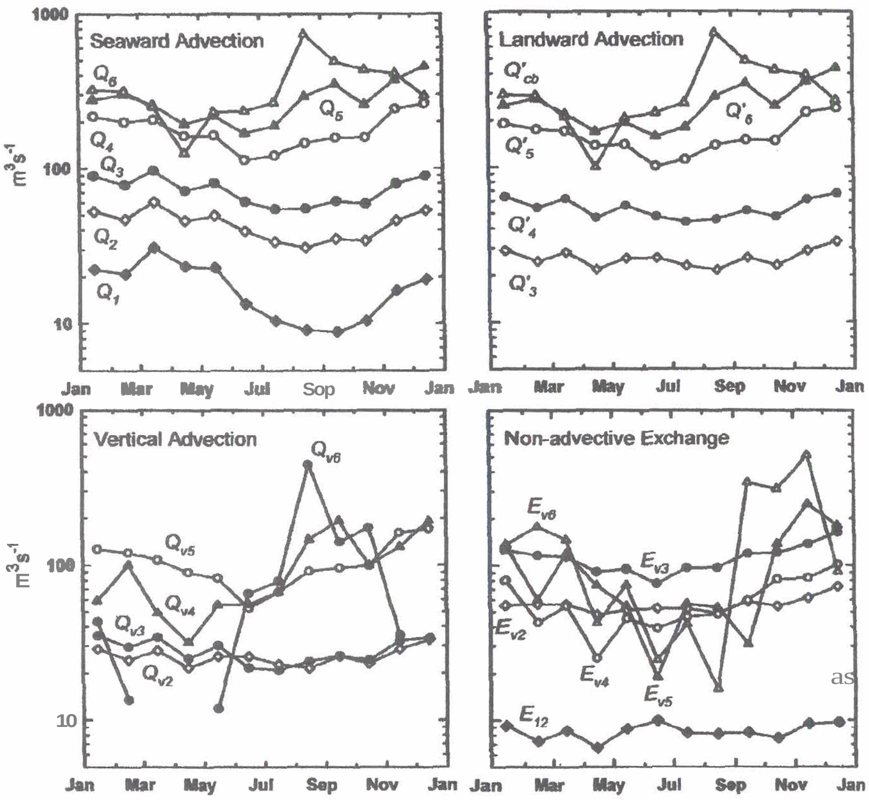

Fig. 4. Exchange coefficients estimated For the 1986–1995 average year. Freshwater inflow to box 1 ($Q_r + Q_{fl}$) is equal to Q_1. March, April and December estimates for Q_{v6} were omitted because they are negative and cannot be plotted on the log scale.

the 1986–1995 average year (Fig. 4) and for each month from 1986–1995. Regular seasonal and spatial patterns were apparent in the seaward advective flows (Q_1, Q_2, ... Q_6), while more varied patterns were evident in vertical advection (Q_{v2}, Q_{v3}, ... Q_{v6}) and non-advective exchanges (E_{12}, E_{v2}, E_{v3}, ... E_{v6}; Fig. 4). Maximum seaward advective flow in the upper and middle estuary (Q_1, Q_2, Q_3, Q_4) occurred during December through March when freshwater inputs were greatest, even though maximum advection in the lower estuary (Q_6) occurred during late summer and early fall. Maximum landward advection into the lower estuary also occurred during late summer and early fall. There was no correlation between landward advection at the estuary mouth (Q'_{cb}) and freshwater inflow. Vertical advection was essentially constant through the year in boxes 2 and 3 (Q_{v2}, Q_{v3}). Maximum vertical advection in box 4 (Q_{v4}) occurred during winter and spring, while in box 6, vertical advection (Q_{v6}) was highest in fall and lower in winter and spring, with negative values ($Q_{v6} < 0$) occurring in April and December. Non-advective vertical exchange coefficients (E_{vm}) were essentially constant throughout the annual cycle in box 2 (E_{v2}). For boxes 3 and 4, non-advective exchange coefficients (E_{v3}, E_{v4}) were gradually decreased into summer and increased again in fall, indicating the effects of strong summer stratification and gradual weakening into fall. Vertical non-advective exchange in boxes 5 and 6 increased abruptly into fall, indicating a more sudden turnover of the water column (Fig. 4). Since vertical gradients in dissolved materials are weakened by this strong vertical mixing, the associated vertical transport, or the product of the exchange coefficient and the concentration difference across the two layers, is not as large as suggested by the coefficient alone.

Residence Times

Specific box residence times increased approximately linearly with distance from the estuary mouth (Fig. 5). The median residence time for freshwater (PRT_F) was 68 d. The median residence time for the most seaward box was only 6 d. The median estuarine residence time, or ERT (Miller and MacPherson 1991), was 25 d.

The magnitude of all the residence times that

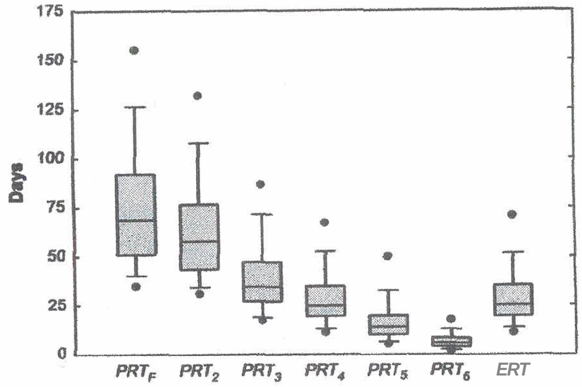

Fig. 5. The distributions of pulsed residence times (*PRT*) for the Patuxent River estuary during 1986–1995. The boxes and whiskers represent the 10th, 25th, 50th, 75th, and 90th percentiles. The black dots show the 5th and 95th percentiles.

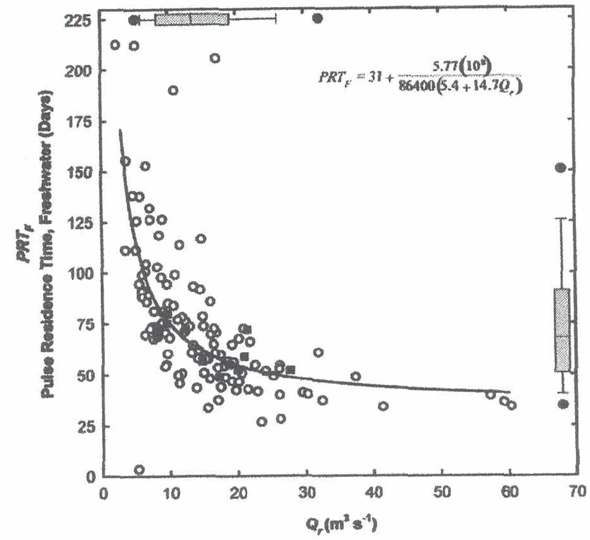

Fig. 6. The relationship between the residence time of freshwater calculated for each month during 1986 to 1995 and freshwater inflow at the head of the estuary. For comparison, the residence times for each month of the 1986–1995 average year (long-term mean months) were also plotted. The regression line was fitted using iteratively re-weighted (robust) regression (SAS Institute, Inc. 1990). A small number of values were well outside the cluster of observations, illustrating the need for either a robust regression procedure of the type that was used or application of an outlier rejection criterion. Box and whisker plots show the distribution of river flow and residence time.

were estimated were strongly predicted by either freshwater inflow (Q_r), seawater inflow at the mouth of the estuary (Q'_{cb}), or a combination of both. Since these two flows were not well correlated, it was possible to observe which residence times depended more on which flow terms. PRT_F was well predicted by Q_r alone (Fig. 6) with the exception of several extreme outliers. *ERT* was better predicted by Q'_{cb} alone (Fig. 7). A multiple nonlinear regression including both flow terms (Eq. 16) illustrated down-estuary changes in the relationship of flushing processes to river flow and seawater inflow (Fig. 8). The top panel in Fig. 8 shows predicted asymptotic residence times as both river flow and seawater inflow become very large. Asymptotic residence time was smaller for down-estuary boxes than for up-estuary boxes. Residence time was well correlated with river flow from the head of the estuary down to box 4, but PRT_5 and PRT_6 were essentially unrelated to river flow (Fig. 8, middle panel). The dependence of flushing processes on seawater inflow increased exponentially as the location of tracer release moved seaward (Fig. 8, bottom panel). Estuarine residence time (*ERT*) depended approximately equally upon river flow and seawater inflow in this analysis.

There were large seasonal differences between freshwater replacement time (*FRT*), the residence time for which steady-state salinity is assumed, and PRT_F, the similarly defined residence time for which this assumption is relaxed. *FRT* was up to 39% higher than PRT_F during June through September, 25–39% less than PRT_F from November through April, and approximately equal in May and October (Fig. 9). Annual means of the two residence times were nearly identical. Seasonal differences between *FRT* and PRT_F were related to,

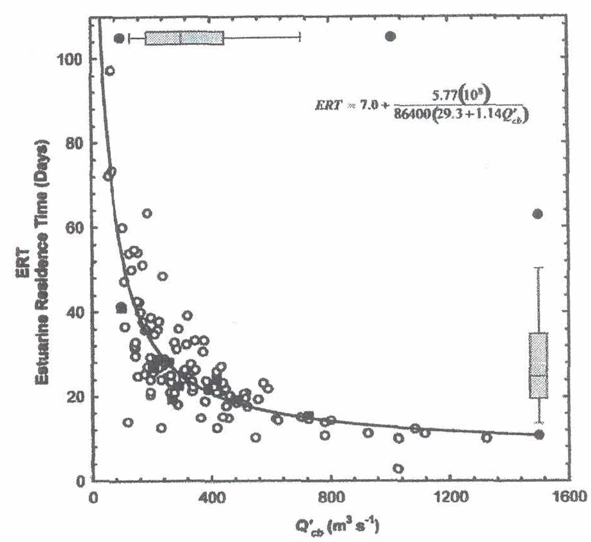

Fig. 7. The relationship between estuarine residence time and saline inflow to the mouth of the estuary calculated for each month during 1986 to 1995. For comparison, each month of the 1986–1995 average year is also plotted. The regression line was fitted non-linear least-squares regression. Box and whisker plots show the distribution of river flow and residence time.

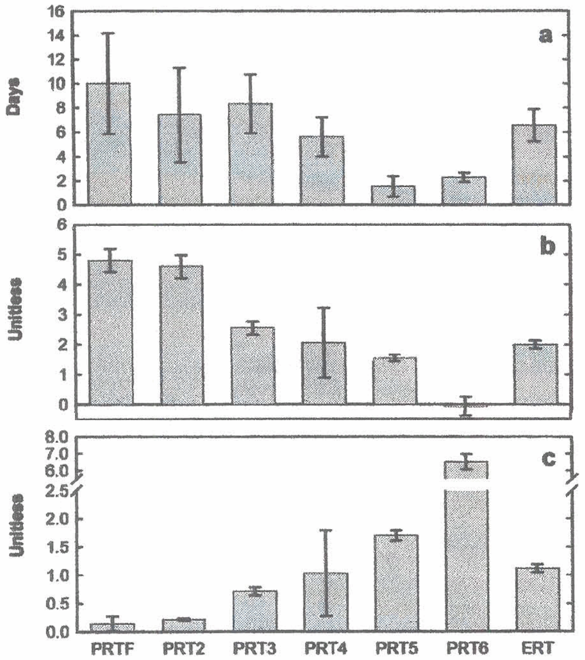

Fig. 8. Model parameter estimates (±SE) for hyperbolic models relating residence times for Patuxent River estuary to freshwater inflow and seawater inflow. The regression models are of the form $T = a + 5.77(10^8) \div 86{,}400(bQ_r + cQ'_{sb})$ where T is residence time in days, Q_r is river flow in $m^3\ s^{-1}$ and Q'_{sb} is the seawater inflow in $m^3\ s^{-1}$. The parameter a has units of days, while b and c are unitless quantities.

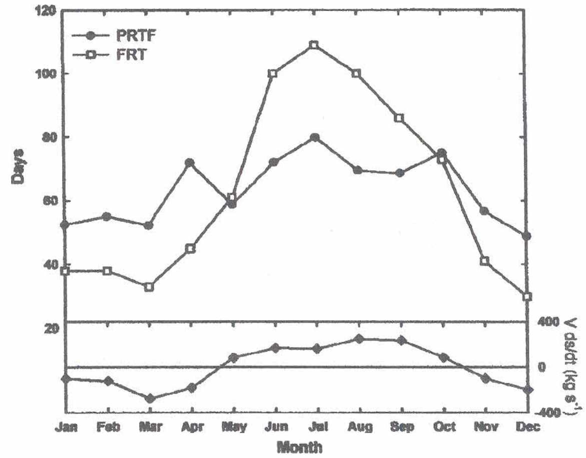

Fig. 9. Pulsed residence times for freshwater (PRT_F) and freshwater replacement times (FRT) for Patuxent River estuary, the latter of which is calculated using the fraction-of-freshwater method (Dyer 1973). The estimates are for the 1986–1995 average year. The rate of change of salt storage in the estuary is plotted below.

but not perfectly predicted by, the seasonal changes in salinity of Patuxent River water. When salinity was increasing, FRT was higher than PRT_F, while the opposite was true when salinity was decreasing. FRT was nearly the same as PRT_F during May and October when salinity changed only slightly (Fig. 9).

Discussion

The box model constructed for the Patuxent River estuary was a simple, effective way to estimate bulk water transport and residence times over seasonal time scales at different times of the year and at different levels of river flow. Relaxing the commonly made assumption that salinity remains constant allowed the model to reveal independent dynamics of river flow and saline inflow at the estuary mouth (Fig. 4). The result wa.. that the model showed that circulation in the lower basin of the estuary and therefore residence time was nearly independent of river flow. This may explain the independence from river flow observed for water quality in the lower basin of the estuary (Hagy 1996).

The detailed hydrological data that were available undoubtedly improved the estimates of freshwater inputs, especially ungauged flows. This was important for the Patuxent River estuary because freshwater inputs below the fall line accounted for 43–61% of the total freshwater inputs (Table 1). The challenge of estimating these flows may be greater in lagoons (e.g., Chincoteague Bay, Maryland; Pritchard 1960) or smaller in estuaries dominated by large rivers (e.g., Columbia River, Washington). Correct accounting for water diversions to municipal water supplies and regulation of flow by dams was important for accurately calculating the runoff rates needed to estimate ungauged flows. For example, the municipal water diversion during August was equal to 39% of the observed fall line discharge, 31% of which was due to dam releases. Smith et al. (1991) found similar artificial controls on the hydrology of the Tomales Bay watershed. Because of the prevalence of water regulation, some attention to detail may be required to construct accurate water budgets for box models.

Direct precipitation and evaporation were a relatively small component of the water budget because the Patuxent River watershed area is 17 times the water surface area (Table 1). Groundwater inputs were not directly estimated, but were implicitly included in the runoff rates because groundwater feeds the gauged portion of the river from which the runoff rates were calculated. For certain other estuaries, estimates of groundwater input rates might be important for constructing a water budget for a box model.

Compared to freshwater inputs, salinity is easily quantified given appropriate data. While data from a single cruise is sufficient to support a rudimentary box model calculation (e.g., Taft et al. 1978), important advantages were gained from the availability of long and detailed salinity records. The spatially dense array of salinity data (Fig. 1) permitted interpolation and therefore more accurate estimation of average salinity. The spatial resolution of the salinity data also permits good spatial resolution in the model. If very high resolution spatial data are available, the spatial resolution of the box model will be limited by the numerical stability of calculations which decreases as the differences between the salinity in adjacent boxes decreases. Good data on the physical dimensions of Patuxent River (Cronin and Pritchard 1975) were needed for proper calculation of mean salinity and for calculating time-variable salinity terms (i.e., Eq. 7). At a minimum, some idea of seasonal changes in salinity is needed to estimate the rate of change of salinity. This study shows that when such changes occur, neglecting them can lead to substantial errors.

The box model reproduced essential features of the 2-layer gravitational circulation typical of coastal plain or drowned river valley estuaries as reviewed by Day et al. (1989). Specifically, an approximately 20-fold down-estuary amplification of seaward advective transport was associated with vertical inputs of water to the surface layer from the landward-flowing bottom layer (Fig. 4). Diffusive exchange along the pycnocline tended to be lower in the summer than in other months, especially in the lower estuary, reflecting strong seasonal stratification (Fig. 4). An encouraging aspect of the model results is the sensible and apparently realistic results obtained from such a simple model. Dividing the surface and bottom layer transport for May-June 1986 by the respective cross-sectional areas yields net current velocities of 5.5 cm s^{-1} and 5.7 cm s^{-1}, respectively. Net non–tidal current velocities of 5.5 cm s^{-1} and 6.0 cm s^{-1} were obtained by time-averaging acoustic Doppler current profiler (ADCP) results obtained for the same period (Boicourt and Sanford 1988). Given the uncertainties inherent in this comparison (see Hagy 1996) and the vastly different methods involved in generating the estimates, this similarity provides a reassuring independent validation of the box model results. This comparison also suggests that our assumption regarding advective transport versus tidal dispersion is not grossly in error, at least for this May–June 1986 period.

Further justification for our neglect of horizontal non-advective exchanges in the middle and lower estuary may be derived by referring to arguments presented in Fischer et al. (1979). They state that non-advective exchange in the direction of a mean flow may be neglected in comparison to advection by the mean flow when the time scale of interest is much longer than $2D/u^2$, where D is the longitudinal dispersion coefficient and u is the mean flow speed. We are aware of no direct estimates of D for the Patuxent River estuary, but Fischer et al. (1979) quote estimates for the adjacent Potomac River estuary from two sources in their Table 7.2. These estimates range between 20–100 $m^2\ s^{-1}$, which, when combined with the above estimate of approximately 0.06 m s^{-1} for u, yield a limiting time scale of 3–15 h. In other words, longitudinal dispersion is likely to be important at and below tidal time scales, but for the monthly time scales of interest here it should be negligible in comparison to advection by the gravitational circulation. Another indication of the relative importance of advective and non-advective exchange is the mass transfer Peclet number, which is the ratio of advective transport to dispersive transport defined in this case by uL/D where L is the longitudinal length scale of interest Using L ≈ 9 km as the average axial distance between–the centers of adjacent boxes and the same estimates of u and D, we obtain Peclet numbers of 5–27, indicating the dominance of horizontal advective exchang due to the 2-layer gravitational circulation in the middle and lower Patuxent River estuary.

An intriguing result of the box model is that the enhancement of gravitational circulation expected when river flow increases was not observed. The landward bottom-layer inflow from Chesapeake Bay (Q_{cb}) was uncorrelated with freshwater inputs. The highest values of Q_{cb} sometimes occurred when Q_r was low (Fig. 4). While the box model is not based on hydrodynamic principles, the results appear to reflect complex estuary-subestuary interactions. An increase in river flow decreases salinity, usually intensifying the salinity gradient and increasing the longitudinal pressure gradient This leads to acceleration of the gravitational circulation until the pressure gradient force is balanced, largely by friction in a small estuary. However, since the Patuxent River estuary is a sub-estuary of Chesapeake Bay, decreases in salinity at the Patuxent River estuary mouth may be caused by increases in Susquehanna River flow. This reduces the salinity gradient across the mouth of Patuxent River. In the upper Chesapeake Bay, such estuary-subestuary interactions are even stronger and can cause reverse estuarine circulation and 3-layer circulation in the tributary estuaries (Chao et al. 1996). Three-layer circulation has been reported at the mouth of Patuxent River by Boicourt and Sanford (1988), mostly during December and April. Since the box model was structured for the more typical 2-layer estuarine circulation, it cannot accurately represent the circulation when these circulation patterns occur. This type of circulation may explain why Q_6 was less than Q_5 (December) and even Q_4 (April) in the average year circulation estimates (Fig. 4). Importantly, if the time-variable salinity terms are neglected, estimated up-bay circulation is proportional to freshwater inputs and the model fails to reflect these more complex circulation patterns.

The residence time for freshwater (PRT_F) had a median value of 68 d, a moderate to long residence time compared to other estuaries. Columbia River, Washington, is flushed in 1 to 2 d due to a very high flow rate (Nixon et al. 1996). Chincoteague Bay, Maryland, is flushed in 10–20 d largely by tidal exchange (Pritchard 1960). In contrast, Chesapeake Bay, Maryland (Nixon et al. 1996) and Guadalupe Estuary, Texas (Longley 1994) have average residence times of about 90 d.

Unfortunately, estuarine residence time is reported much less often than residence time for freshwater because it can only be calculated using a simulation approach. This residence time, which has the useful property of reflecting the flushing rate for the average water parcel in the estuary, had a median value of 25 d for the Patuxent River estuary. The fact that estuarine residence time was best predicted by Q'_{cb} rather than Q_r (Fig. 7) indicates that for much of Patuxent River estuary, factors affecting the two-way exchange with Chesapeake Bay determined flushing rates more than freshwater flow. This implies that water quality effects associated with nutrient enrichment in high flow years may not be offset by greater flushing rates, offering an explanation for the unusually good, but region-specific, relationships observed between middle Patuxent River estuary water quality and Patuxent River flow (Hagy 1996).

Other residence times besides that for freshwater reflect the different periods of retention for water parcels originating at different points in the estuary. Residence times decre ed as the tracer origin was moved closer to the mouth of the estuary, as implied by the definitions used. If PRT_F barely exceeds PRT_4, a rapidly flushed upper estuary is indicated but if PRT_4 barely exceeds PRT_6 and PRT_F greatly exceeds PRT_4, a slowly flushed upper estuary and a rapidly flushed lower estuary is indicated. These differences can have important effects on water quality distributions (Hagy 1996). The models show that these differences arise from the lack of correlation between river flow and gravitation circulation.

The down-estuary increase in the importance of saline inflow relative to Patuxent River flow (Fig. 8), combined with the lack of correlation of these circulation components creates seasonally and spatially varying patterns of flushing in the Patuxent River estuary. Given the seasonal differences in the relative magnitudes of river flow and saline inflow, a seasonal alternation of likely retention zones in the estuary occurs. Specifically, particles and dissolved materials are more likely to be retained in the upper estuary during summer and fall when river flow is low and in the lower basin when exchange with Chesapeake Bay is minimal during the spring. Although confounded by biological processes, these patterns are reflected dramatically in water quality patterns (Hagy 1996), suggesting important biological-physical couplings.

Summary

Box models are an effective way to estimate bulk physical exchanges and residence times using frequently available or easily collected data sets. The model equations can be solved only if the series of salt and water balance equations can be reduced to an equal number of equations and unknown quantities. This is simple for an unstratified estuary, but required a simplifying assumption in this case. Incorporation of additional terms in the model equations reflecting the time-rate-of-change of salinity was an important improvement to the box modeling methodology that enabled season-specific circulation estimates to be made where steady-state salinity cannot be assumed. Good estimates of a variety of residence times can be easily calculated using transport estimates obtained from such box model simulations. For the Patuxent River estuary, these estimates were well predicted by river flow, saline inflow, or both. Changes in the relative importance of these two factors in different regions of the estuary indicated an important feature of the circulation of Patuxent River estuary that appear to be reflected in its ecological responses to nutrient enrichment.

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